MAXIMUM INTENSITY OF GROUND MOVEMENTS CAUSED BY FAULTING

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Abstract. In this paper it is shown that a local fracture in a rock mass or slip on an existing fault will release an amount of strain energy which will be some fraction of the total energy recoverable from that stored in the mass at breaking point. And since this is a function of the properties of the strained material, an upper limit for the velocity of fault movement at fracture can be established, which will be of the order of 100 cm/sec. It is also shown that the residual strength on a fault plane should be small, not greater than 100 kg/cm^2, and that the strength drop at failure should also be small.

Introduction. The major sources of energy capable of producing ground movements of engineering interest are due to blasting and earthquakes. Since the former arise from the use of either chemical or nuclear explosives, the energy input at the source is usually controllable. In contrast, earthquakes develop in response to tectonic agencies and the energy released at the source cannot be predicted at the outset.

The majority of hypotheses for the generic cause of shallow earthquakes are based on the sudden release of strain energy caused by fracture or slip along existing faults. The observed pattern of the direction of the onsets of certain waves at the surface and the relation among their amplitudes suggest a source composed of a transient force or displacement system applied at the focus. There is, naturally, a number of cases where earthquake mechanisms based on single or double couple of forces are not applicable, the available evidence suggesting instead a mechanism of focal volume change (13,35).

There may well be a number of mechanisms which, either together or separately, can produce an earthquake; but any such mechanism will require a transient change in strain energy over a certain volume in the earth's crust or upper mantle. Two of these mechanisms are, of partial elastic rebound and of polymorphic transition (6), the former predominantly confined in the crust and the latter deeper down.

The rebound mechanism, being associated with ground surface effects, is of great interest to the engineer. This mechanism requires a medium capable of storing strain energy and also releasing part of it transiently. The crustal masses have had a complicated stress history and apart from the strain energy that they may contain due to gravity and tectonic forces, they may also contain some due to residual stresses (9,11). A further increase in stress or a decrease in strength in some part of the mass may bring about instability and local failure of the material. As a consequence of this, part of the stored energy will be released suddenly and will propagate in the form of waves. A stress increase can be brought about by increasing or by decreasing overburden pressures and local stresses, natural or man-made, such as by deep excavations (9), mining and tunneling works, rapid accumulation of thick deposits, impounding of reservoirs (8,14,31,36), and by stress waves caused by explosions.

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Stress changes can also be due to tectonic forces and by the rock expansion caused by alteration (21). A decrease in strength can be brought about by changes in the effective stresses due to abnormal increase in the pore fluid pressure of the material, by weathering and alteration (12).

Thus, the two main sources of vibrational energy, explosions and earthquakes, are respectively "sources" and "sinks" of strain energy. Explosions will in general produce a sudden increase in the total energy content of the medium and the flux of energy at any distance from such a "source" can be predicted. In contrast with explosions, where the energy flux at the source can be made many times greater than that required to cause fracture, the flux at the focal volume of an earthquake should be limited. This follows from the fact that a local fracture in a rock mass or slip on an existing fault will act as a "sink" that will drain an amount of strain energy which will be some fraction of the total energy recoverable from that stored in the mass at breaking point. And since this in turn is a function of the properties and volume of the strained material, an upper bound to the amount of energy that can be stored, and hence that can be released by fracture per unit volume of strained material, may be established. Various estimates of the seismic energy per unit volume of the mass from which the major part of the energy is released, show roughly $10^{2.5}$ to $10^{3.0}$ c.g.s, a figure independent of the total energy content at the source (3,7,37).

In the case of localised or contained fracture within a mass, the kinematic constraints that will prevent fracture from propagating will allow only partial stress release and this will leave residual stresses. These stresses, together with the fact that the residual strength of a large rock mass will depend on the ambient effective stresses and also that this strength after failure will have a finite value, exclude total stress relief on the fault after fracture or slip.

Partial Elastic Rebound. Earlier on we pointed out that the strain energy released by fracture will be some fraction of the total energy recoverable from that stored in the mass at breaking point. Assuming that fracture is associated with a sudden movement of a fault in the crust, we can show that the maximum velocity of fault movement should not exceed 100 to 150 cm sec.

Consider an infinitely long, vertical plane failure surface with an average shear strength $P/2$, extending through an elastic solid of shear modulus G and density ρ, Figure 1. Moreover, assume that the solid is under simple shear, strained slowly parallel to the vertical plane, thus storing uniformly an increasing amount of potential energy. At a certain moment, the shear stress on the plane will reach the peak shear strength of the material and the two sides of the solid will tend to slip in opposite direction. Let $pP/2$ be the value of the average residual strength of the fault plane after failure, where p is less than unity, Figure 2a, and assume that during slipping the resistance on the fault plane will be given by $(pP/2 - 2ku)$, where u is the displacement of the fault and k is a parameter proportional to the rate at which the residual strength on the plane decreases with relative displacement. This parameter can be used to introduce the effect of pore fluid pressure or temperature on the residual strength of ~~the~~ fault. It can be shown that for this model of faulting, the velocity with which the two sides of the fault will move after failure is:

$$v = (2PS/G)\left[(kh/G) + (1-p)\right]\sum_n a_n \sin(a_n St/h)\cdot\left[(kh/G) + (kh/G)^2 + a_n^2\right]^{-1} \quad \ldots(1)$$

where a_n are the roots of $(kh/G)\tan(a_n) + a_n = 0$, and S is the shear wave velocity in the medium.

At present, we have no data on the basis of which to calculate k, but from equation (1) it appears that the velocity of slip v will tend to increase with increasing values of k. For $k = 0$, the maximum velocity with which the fault will move can be obtained from equation (1) in a closed form:

$$v = PS(1 - p)/2G \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots(2)$$

When $p = 1$, that is when the drop in strength on the fault is zero, there will be no shock, and the velocity will be zero. The maximum velocity of fault movement obtains when $p = 0$, i.e. when the strength on the fault plane drops to zero, in which case equation (2) reduces to the expression derived by Housner (19).

Thus, the velocity of fault movement will depend on the strength drop parameter $(1 - p)$. Laboratory tests on rock specimens under low to medium ambient pressures show that the strength at failure is in general larger than the strength that is available after failure at large displacements (5,15,16, 17,29,34). For the more brittle materials the transition after failure is very rapid with a sudden drop in strength, a process that is obviously undrained. This drop in strength seems to reflect a partial or total loss of cohesion rather than a sudden reduction in friction on the fault. Experimental evidence from small rock specimens shows that the strength drop parameter $(1 - p)$ decreases with increasing effective confining pressure or depth, and that above a certain pressure the behaviour of the rock becomes ductile, Figure 2b. Thus, other things being equal, shallow faults will show, in principle, larger velocities of slip than deeper ones. Also, the stress relief on a shallow fault will, in general, be larger than in a deeper fault.

The velocity of fault movement will also depend on the value of (P/G) which, for a material in a brittle state is a measure of the critical strain at failure. This strain is not the observable strain on the surface of the earth after faulting but the strain from a position of equilibrium. What we usually measure after faulting is in fact a strain change, total relaxation of which is impeded by kinematic constraints and by the residual strength on the fault. Values of (P/G) are not really known for actual conditions. The very existance of a fault in nature, however, is evidence that the crust must have been strained a number of times in the past beyond its purely elastic critical strain, so that subsequent critical strains should depend on the residual strength on the fault rather, with values of the order of 10^{-3}.

Thus, for a shallow fault, extending down to 20 kilometres, with critical strain, say $3x10^{-3}$, and shear wave velocity of 3 km/sec, a strength drop at failure of 25%, will be associated with a velocity of fault movement of about 120 cm/sec. This would be almost an extreme value for the velocity because our model entails elastic behaviour of the material, an infinitely long fault plane and ignores the finite dimensions of the fault as well as the stresses that will result by local rotations and dilatations of the crust at the time of rupture. The value of the stress drop used here should also be an extreme as it is based solely on laboratory results on very small specimens. The strength measured in the laboratory should be larger than that which obtains in a large mass or on an old fault zone before rupture. There must be a scale effect which will depend on the size of the strained volume. As we pass from a local fracture in a small rock sample to an actual fault in nature and the size of the strained volume increases, it becomes increasingly probable to find flaws and high local stresses that will result in an average strength much smaller than that which obtains in the laboratory. The same scale effect must apply to the critical strain (P/G) which on a fault will be smaller than that which obtains in the laboratory from intact small specimens. Moreover, the frictional stress that may oppose shear motion on an existing fault should be much smaller than the normal stress on the plane. The 45-degrees angle of friction postulated for fault movement, in nature, should be an unrealistic figure.

Another line of argument that suggests an upper bound for the velocities of fault movement is that of the apparent absence of melting on fault surfaces. Most exposed faults give little or no evidence that vitreous or amorphous materials, such as pseudotrachyte, have formed on their sides (see for instance 22, 39). This implies that not only the relative velocity of slip should be comparatively small but also that the residual strength pP/2 on the fault should be far smaller than that obtained in the laboratory from unfractured rock specimens. It is not difficult to show for our simple model, shown in Figure 1, that the temperature rise above zero on the fault during slip is given by

$$\theta = pPQv(t)^{\frac{1}{2}}/KJ(\pi)^{\frac{1}{2}} \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad (3)$$

where K and Q^2 are the conductivity and diffusivity of the rock respectively. In equation (3), v is the velocity of slip, and J is the mechanical equivalent of heat.

Before we attempt to evaluate the probable temperature rise on a fault surface, let us try to assess the order of magnitude of the residual strength that obtains during actual faulting. If L and h are the average length and depth of faulting, the work done in shear over a relative displacement R of the two sides of the fault would be (pPLhR). This work will be some fraction of the total energy release in the form of seismic waves (qE), where E is the seismic energy and q is the efficiency of the shock. Thus, as a first approximation (pP) can be obtained from

$$pP = (qE)(LRh)^{-n} \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad (4)$$

Table I gives a list of earthquakes after 1900 which are associated with faulting. Sixty-two cases are included in this list, seventeen of which are still under study[1]. In compiling this table I resorted to the original sources of information and avoided using data tabulated at second or third hand. Relative displacements R refer to the resultant of the maximum horizontal and vertical components of movement. For a number of cases it was possible to calculate the average resultant displacement R_{av} which was found to be 1.5 to 10.0 times smaller than the maximum displacement R.

Figure 3 shows a plot of log(qE) versus $\log(LR_{av}h)$. Here, we assumed that q = 1 and h = 10 km, while E was calculated from Båth's magnitude-energy relation $\log(E) = 12.2 + 1.44(M)$, where M is the magnitude.

From the cases for which we have values of R_{av}, we find that in equation (4), n = .1 and (pP) = o.1 kb. These values and their mean will tend to increase slightly if we include all cases listed in Table I assuming $R = 3R_{av}$.

If we consider now that the displacements will be zero along the vertical ends and lower part of the fault plane, the value of (pP) found earlier should be increased by a factor of $(\pi)^2$. On the other hand, if the efficiency (q) is taken to be 10.0, the net effect on the value of (pP) = o.1 kb, is negligeable.

Thus, we find that on the average, the residual strength on a fault plane will be of the order of 100 kg/cm^2 rather than of the order of kilobars. This strength will be purely frictional, and on the whole its value is not inconsistent with strengths calculated from other considerations (4,23). The strength drop then, required to cause a "sink" of energy should correspond to the cohesive strength of the fault before fracture and to the strength contributed by kinematic constraints. The ratio of this strength to the total strength should decrease with increasing ambient effective stress, or depth, an observation concurring with what we said earlier on the variation of (1 – p) with depth.

We may now return to equation (3) and calculate the temperature rise on a fault-break. Taking a residual strength pP = o.1 kb, with K = o.06 and Q^2= o.08 c.g.s., we find that in order to have a temperature on the fault below melting point, say 1,000 °C, a relative displacement of 50 cm. cannot take place with velocities greater than about 80 cm/sec. If the residual strength is taken at 1.0 kb, a 50-centimetre movement will cause melting if it occurs with a velocity greater than 1 cm/sec.

Although there are many questionable assumptions in these simple engineering calculations, it appears very probable that the maximum ground velocities near a fault-break should not exceed 100 to 150 cm/sec. More exact values of these velocities will require more information on the residual strength and of the actual movements during faulting.

In contrast with ground velocities, it is not possible to deduce an upper limit for the ground accelerations. Formally, these can be quite large and there is no good reason why ground accelerations locally should not exceed values of 100%g. These large accelerations, however, are of little importance to the

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(41): Ambraseys N., Tchalenko J. (1968) "Documentation of faulting associated with earthquakes" UMP Report.

engineer since they will be associated with very high frequencies.

<u>Field evidence of strong ground movements</u>. The most commonly used method for the indirect measurement of the ground acceleration has been to calculate the inertia force required to overturn rigid, free-standing bodies. Refinements of this method include the rocking effect of the overturning force and the effect of vertical acceleration.

For earthquakes of a shock type, of short duration, maximum horizontal and vertical accelerations have been determined from a large number of measurements on objects that slid during the earthquake, leaving a straight, visible trace of their movement. In this method, the length of the trace and the coefficient of friction between the object and the ground can be measured. The unknowns then are, the shape of the horizontal and vertical acceleration pulses, their amplitudes and duration. For a sufficiently large number of cases, these unknowns can be determined by least square fitting of the observed and measured quantities to the appropriate equations of motion (Appendix).

All these indirect calculations give quite high values for the ground acceleration, in cases approaching 100%g, but anomalies have been found, and until these are explained it is not easy to say to what extent the results from such calculations can be trusted. Recorded maximum accelerations are also large. Before 1966 the largest acceleration recorded was 33%g; in 1966 an acceleration of just over 50%g was recorded at Parkfield only to be exceeded a year later at Koyna by a value of 62%g (2).

Formally, there is no upper bound for the maximum ground acceleration and provided recording instruments will be capable of recording faster accelerations of higher frequency than today, there is no reason why accelerations exceeding 100%g should not be recorded in the future. At fracture near a fault, the rate of velocity change will depend on the mode and linear dimensions of the rupture. For instance, in rock bursts predominant frequencies are in the range of o.1 to a few kilocycles, and in rock specimens tested in the laboratory predominant frequencies are even higher (27). The wave lengths of the waves emitted by the formation of new cracks in some cases are of the same order of magnitude as the length of the cracks, and in earthquakes associated with surface faulting these frequencies may well vary from a fraction of a cycle to tens of cycles per second (32). Thus, the rate at which a velocity, otherwise small, will reach its maximum value may be quite high.

Numerous attempts have been made in the past to correlate maximum ground acceleration with local Intensity, and various correlation formulae are now available. Table 2 shows the results of the analysis of 33 strong motion records and Figure 4 shows a plot of the maximum recorded acceleration versus Intensity (MM) at the recording station. From this Figure we notice that accelerations recorded for a particular Intensity rannge over two orders of magnitude and that the scatter is too large to allow a correlation between intensity and acceleration or, to justify the usefulness of current acceleration-Intensity conversion formulae.

The question now arises whether the maximum recorded acceleration is in fact maximum. At any given instant, the maximum ground acceleration will be larger than the maximum value of either of the three recorded components. Combining at intervals of about 10 milliseconds the two horizontal components of the records listed in Table 2, we find that the maximum resultant acceleration is on the average larger by 5% than the maximum component. The maximum resultant velocity was found to be about 15% larger than the maximum component. Under favourable, though extremely improbable conditions, this difference should be 42%. Thus, the maximum recorded acceleration is not much different from the true spatial maximum.

From Table 2 we notice that maximum ground velocities do not correlate with Intensity either. These velocities have been obtained through integration of the corresponding acceleration records and they correlate very strongly with the energy flux at the station. The total energy at the station, including both kinetic and potential energy, on a record will be $E = 2(E_b + E_s)$. The total kinetic energy E_b of spherically diverging body waves, measured on the surface of the ground can be calculated approximately by

$$E_b = A\sum_i v_i^2 \, \Delta t_i S_b \quad \ldots\ldots\ldots\ldots\ldots (5)$$

Assuming that the surface waves spread radially in a cylinder with a vertical axis, the total kinetic energy E_s is given approximately by

$$E_s = B\sum_i v_i^2 \, \Delta t_i S_s \quad \ldots\ldots\ldots\ldots\ldots (6)$$

where $A = \pi R^2 \rho/2$,

$B = \pi R h \rho$ and R is the focal distance of the station, ρ is the mass density, h is the focal depth, and S_b and S_s are the velocities of propagation; v_i is the average particle velocity between two successive crossings of the zero position at the trace, and Δt_i is the time of the interval, all in c.g.s units.

We may now define the energy flux at the station as $F = E/A_s t_o$, where A_s is the surface of the sphere that passes through the station with centre at the focus and t_o is the duration of the significant part of the record. The flux now becomes the average energy density at the station (33), which shows an excellent correlation with the maximum ground velocity. From the records listed in Table 2, we find that the following relation between F and v holds

$$\log(F) = \log(v^2) + 3.1 \quad \text{(c.g.s.)} \quad \ldots\ldots\ldots\ldots\ldots (7)$$

In deriving equation (7) we assumed spherical radiation for all cases in Table 2, except for the events no.8, 18, 19, 20 and 21. For event 8 we assumed cylindrical and for the rest, plane surface radiation. The duration t_o was taken arbitrarily equal to the time during which the ground accelerations were equal to or greater than 3%g.

We thus see that the energy flux and the maximum particle velocity not only correlate with each other, but also that the constants in equation (7) have reasonable values. In the first place, the velocity, as it should, appears in the second power, while the constant 3.1 implies that for a mass density of say 2.5 c.g.s, the average velocity of propagation is 5 kilometres, Figure 5.

Next we notice that equation (7) can be used to deduce an upper bound for the particle velocity within the focal volume. The seismic energy per unit volume V of the mass from which the major part of the energy is released, is more or less a constant and independent of the total energy content at the source, i.e $E/V = 10^{2.6}$ to $10^{3.0}$ (3,7). Assuming a spherical volume of radius R, this ratio can be written as $E/V = 3E/A_s R$. For aspeed of faulting of say 3×10^5, the critical duration t_o may be taken equal to $2R/3 \times 10^5$. Thus, the flux within the focal volume will be of the order of $F = 2 \times 10^7$ c.g.s. Introducing this value into equation (7) we find a velocity $v = 100$ cm/sec; or $v = 130$ cm/sec for $\log(E/V) = 3$.

These values are not much different from those derived using equations (2) and (4). The two methods of calculating v rest on completely different principles and data, and it is remarkable that the results are so similar.

Conclusions. The foregoing suggests that there is an upper limit for the maximum ground velocity of fault movement and that this velocity should be of the order of 100 cm/sec. The ground accelerations, however, can be very large but of little importance since these will be associated with high frequencies. There is no relation between ground acceleration or velocity and local Intensity. The maximum accelerations and velocities of ground movement as measured from individual strong motion records are by 5% and 15% smaller than the resultant movements. There is some indication that the residual strength on existing faults after rupture should be small, of the order of 100 kg/cm², and that the frictional stress that may oppose motion on an existing fault must be smaller than the normal stress on the fault plane.

Appendix. A rigid body of weight W subjected to a sudden movement of its base, may slide on it, overturn, rock or remain stationary. The particular response of the body will depend on the type and duration of the movement, and also on the way in which the body is bonded to the base. Here we consider the case of sliding.

Let $k_o g$ be the maximum horizontal acceleration of the base and f_o the coefficient of friction between the body and the base which is inclined to the horizontal by an angle ϕ. The base is assumed to be plane and the cohesive force between the body and the base to be F. If ag is the vertical acceleration of the inclined base and

$$k = k_o(\cos\phi + f_o \sin\phi)$$

$$f = (F/W) + f_o(1 \pm a)\cos\phi + (1 \pm a)\sin\phi \qquad \text{with} \quad n = k/f, \quad \text{for}$$

$n > 1$, a square pulse of duration T will cause a displacement

$$u = 1/2\ kT^2(1/n - 1) \qquad \ldots\ldots\ldots\ldots\ldots\ldots\ldots(8)$$

An acceleration pulse of <u>half-sine</u> of amplitude k_o and duration T will induce displacement

$$u = (kT^2/n\ \ ^2)(1 - \cos r - 1/2\ r^2n^2) \qquad \ldots\ldots\ldots\ldots\ldots\ldots\ldots(9)$$

if T_m T with $r = \sin r + (1 - \cos r)(n^{-2} - 1)^{1/2}$ or,

$$u = (kT^2/n\ \ ^2)\ \ n^2 + 1/2 \cdot 1 + (1 - n^2)^{1/2}\ \ ^2 - n(\ \ - c) \qquad \ldots\ldots(10)$$

if T T with $n = \sin(c)$, where T_m is the time after the beginning of the pulse at which sliding of the block will stop.

An acceleration pulse of <u>triangular</u> shape of amplitude k_o, duration T, and maximum amplitude at t = mT T, will cause displacements

$$u = (kT^2/6)\ \ 4(1-m)(1-mn) - (1-mn^2)^2\ \ /4n \qquad \ldots\ldots\ldots\ldots\ldots\ldots\ (11)$$

for T_m T, and

$$u = (kT^2/6)\ (1 - n)^3\ \ 2 + 2\ (1 - m)^{1/2} - m \qquad \ldots\ldots\ldots\ldots\ldots\ldots\ (12)$$

for T_m T.

Figure 6 shows the variation of the dimmensionless displacement (u/kT^2). with $p = n^{-1}$, for thethree different pulses considered. The vertical acceleration ag has been considered to act on the base for as long as motion continues. For a brittle bonding between the body and the plane, movement will require n 1 with $F \neq 0$. However, after movement commences, F should be put equal to zero, (1,26).

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TABLE I

	Date	Location	M	L	R	R_{av}
1	1905 Jul. 9	Mongolia	8.3	(115)	(100)	--
2	1905 Jul. 23	Mongolia	8.2	315	540	--
3	1906 Mar. 17	Formosa	7.1	(14)+	310	--
4	1906 Apr. 18	San Francisco USA.	8.3	430	650	(290)
5	1911 Jan. 3	Tien-Shan	8.4	(310)	900	--
6	1912 Aug. 9	Turkey	7.7	(3)+	(250)	--
7	1912 Nov. 19	Mexico	7.0	(100)	(60)	--
8	1915 Oct. 3	Pleasant Valley USA	7.6	65	490	290
9	1925 May 23	N. Tajima , Japan	7.0	(7)+	100	--
10	1927 Mar 7	Tango, Japan	7.5	(70)+	365	74
11	1927 May 22	China	8.2	(60)+	--	--
12	1928 Jan. 6	Kenya	7.0	32	330	100
13	1928 Apr. 14	Bulgaria	6.8	38	40	28
14	1928 Apr. 18	Bulgaria	7.0	62	100	45
15	1929 Jun, 16	New Zealand	7.8	(30)+	450	--
16	1930 Nov. 25	Idu, Japan	7.1	32	360	95
17	1931 Feb. 2	New Zealand	7.8	(8)+	200	--
18	1932 Sep. 15	New Zealand	6.3	(5)+	--	--
19	1932 Sep. 26	Greece	6.9	(6)+	400	--
20	1932 Dec. 21	Cedar Mts. USA	7.2	61	105	20
21	1934 Jan. 30	Excelsior Mnts. USA	6.3	(2)+	13	--
22	1934 Mar. 12	Kosmo, USA	6.6	(10)	50	--
23	1935 Apr. 20	Formosa	7.0	(25)	300	--
24	1935 May 30	Quetta, Pakistan	7.6	(34)	--	--
25	1938 Apr. 19	Turkey	6.7	(15)	115	--
26	1939 Dec. 26	Turkey	8.0	350	420	190
27	1940 May 18	Imperial Valley, USA	7.1	(60)	580	170
28	1942 Jun, 24	New Zealand	7.2	(1)+	90	--
29	1942 Dec. 20	Turkey	7.3	34	200	67
30	1943 Sep. 10	Tottori, Japan	7.3	(44)	(150)	--
31	1943 Nov. 26	Turkey	7.6	265	150	57
32	1944 Jan. 15	Argentina	7.6	(75)	70	--
33	1944 Feb. 1	Turkey	7.6	190	360	180
34	1945 Jan. 12	Mikawa, Japan	7.0	(9)+	240	130
35	1946 Nov. 10	Peru	7.2	(20)+	350	--
36	1947 Apr. 10	Manix, USA	6.4	(4)	6	--
37	1948 Jun. 28	Kukui, Japan	7.2	(45)	180	--
38	1950 May 21	Peru	6.0	(5)+	100	--
39	1950 Dec. 14	Fort Sage, USA	5.6	9	20	5
40	1951 Jan. 24	Superstition Mnts. USA	5.6	(3)	--	--
41	1951 Oct. 21	Formosa	7.1	(8)	230	--
42	1951 Nov. 24	Formosa	7.3	40	210	--
43	1952 Jul. 21	Kern County, USA	7.8	70	170	50
44	1952 Oct. 8	China	5.3	--	--	--
45	1953 Mar. 18	Turkey	7.6	58	430	210
46	1945 Feb. 11	China	7.2	(40)	--	--
47	1954 Jul. 6	Fallon, USA	6.6	20	30	20
48	1954 Aug. 24	Stillwater, USA	6.8	35	75	34
49	1954 Dec. 16	Fairview, USA	7.4	53	800	210
50	1954 Dec. 16	Dixie, USA	7.1	40	460	130

51	1956 Feb. 9	San Miguel, Mexico	6.8	20	120	50
52	1957 May 26	Turkey	7.1	40	160	--
53	1957 Dec. 4	Mongolia	8.3	275	1280	455
54	1959 Jan. 31	Tesikaga, Japan	6.2	(2)	10	--
55	1959 Aug. 18	Hebgen, USA	7.1	30	610	370
56	1961 Jun. 1	Wollo, Ethiopia	6.4	(40)	190	--
57	1962 Sep. 1	Iran	7.2	100	90	30
58	1964 Oct. 6	Turkey	6.8	(30)	--	--
59	1966 Jun. 28	Parkfield, USA	5.5	38	5	2
60	1966 Aug. 19	Turkey	6.8	(30)	--	--
61	1966 Oct. 9	Sudam	5.7	(10)	(10)	--
62	1967 Jul. 22	Turkey	7.2	80	230	60

Note : L = length of fault-break in kilometers.

(L)= Length imperfectly known; (L) + = Fault-break extending into sea or into undocumented region.

R = Relative displacement of fault-break in centimetres; maximum resultai displacement of horizontal and vertical components.

(R)= Imperfectly known; small number of measurement points.

R_{av}= verage resultant displacement in centimetres along fault-break.

(Data based on Ref. 41).

TABLE II

	Date	Station	M	h (km)	D (km)	I	a (%g)	v (cm/sec)	t_o (sec)
1	1952 Jul. 21	Taft	7.7	20	(70)	VII	18	20.0	37
2	1964 Jun. 16	Niigata	7.5	40	38	VI+	16	(60.0)	20
3	1966 Oct. 17	Lima	7.5	30	180	VII	39	20.0	40
4	1962 May 19	Mexico	7.1	30	240	VI	3	1.6	38
5	1949 Apr. 13	Olympia	7.1	70	40	VIII	32	23.0	35
6	1940 May 19	El Centro	6.7	24	20	VIII	33	43.0	30
7	1954 Dec. 21	Eureka	6.6	40	23	VII	26	33.5	10
8		Ferndale			42	VII	20	38.0	20
9	1965 Apr. 29	Olympia	6.5	55	50	VII	19	16.0	18
10		Seattle			25	VIII	8	17.0	16
11	1934 Dec. 30	El Centro	6.6.	15	32	VIII	26	28.0	25
12	1964 Dec. 10	Koyna	6.5	(20)	(30)	VIII	62	25.3	16
13	1941 Oct. 3	Ferndale	6.4	(25)	80	VI	12	7.1	12
14	1933 Mar. 11	Vernon	6.3	10	55	VII	19	20.0	13
15	1941 Jul. 1	S. Barabara	5.9	3o	25	VII	24	20.3	8
16	1951 Oct. 8	Ferndale	5.6	22	55	V	12	7.6	10
17	1938 Sep. 12	Ferndale	5.5	13	55	VI	16	7.6	12
18	1966 Jun. 28	Parkfield-2	5.5.	5	0.1	VII	50	72.2	18
19		Parkfield-5			5	VII	47	27.3	17
20		Temblor			6	VII	41	21.0	7
21		Parkfield-8			9	VII	28	12.6	12
22		Parkfield-12			15	VII	7	6.5	8
23	1957 Mar. 22	Golden Gate	5.3	(10)	13	VI	13	6.1	4

24	1949	Mar.	9	Holister	5.3	12	16	VII	23	12.1	9
25	1966	Apr.	5	Matsushiro	5.1	4	4	VI	42	--	4
26	1957	Mar.	18	Port Hueneme	4.7	5	7	VI	17	12.4	1
27	1966	May	2	Matsushiro	4.7	4	5	V	29	--	1
28	1966	Aug.	8	Matsushiro	4.7	5	2	(V)	39	--	3
29	1966	May	28	Matsushiro	4.7	3	2	VI	37	--	7
30	1966	Jun.	28	Matsushiro	4.6	5	3	(V)	28	--	2
31	1966	Aug.	20	Matsushiro	4.4	3	6	V	30	--	5
32	1966	Aug.	29	Matsushiro	4.3	3	4	(V)	35	--	2
33	1966	Jun.	19	Matsushiro	3.9	5	3	(V)	30	--	2

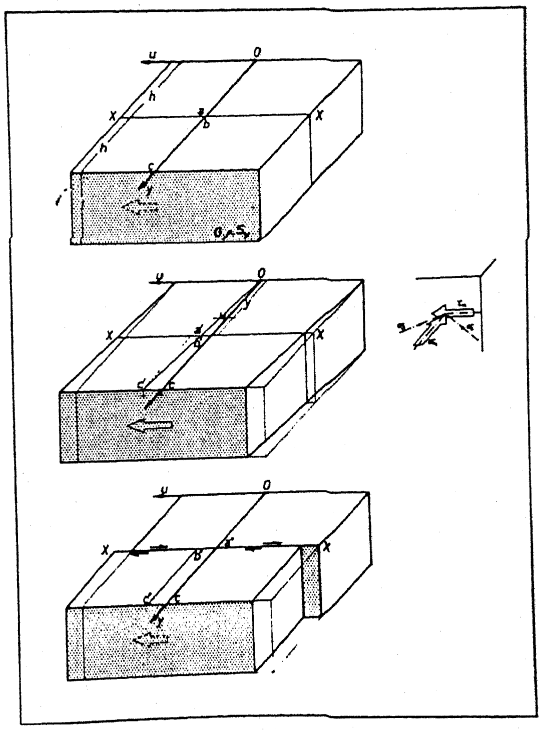

Figure 1.

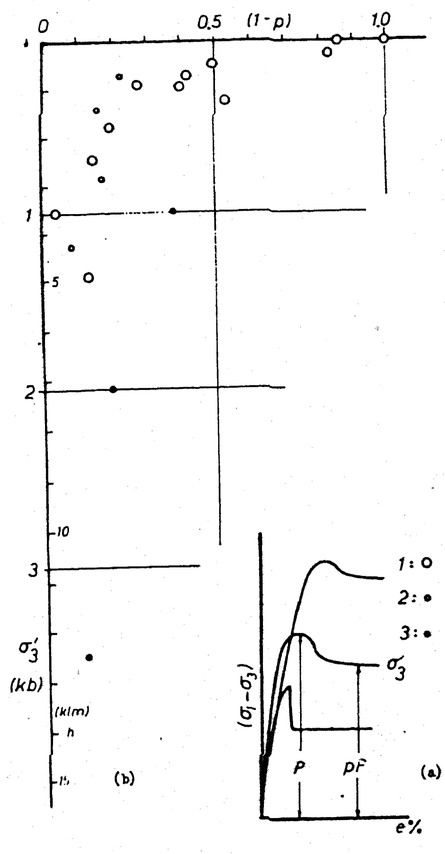

Figure 2. Variation of strength drop with ambient pressure.
1: Ref. 29; 2: Ref. 10, 3: Ref. 34

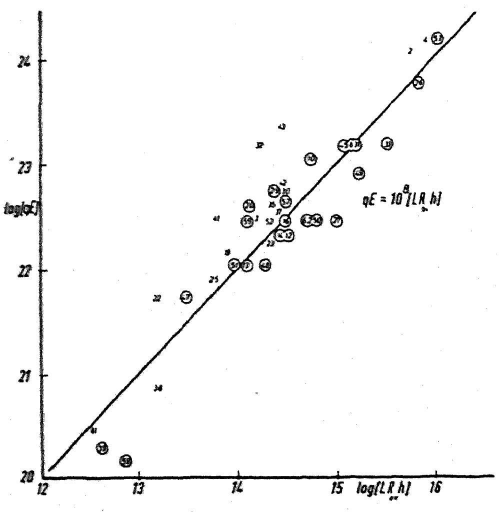

Figure 3
(Numbers refer to Table 1)

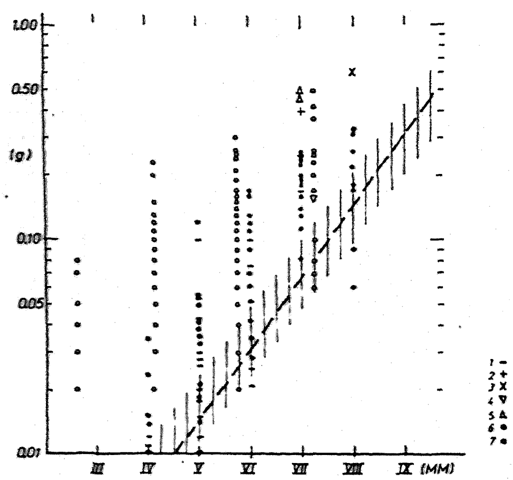

Figure 4. Intensity – Instrumental acceleration plot.
1:18, 2:28, 3:2, 4:30, 5:20, 6:40, 7:24,25
(Numbers refer to source of information).
——— $\log(a) = \frac{1}{3}I_o - \frac{1}{2}$, ///// Ref.38

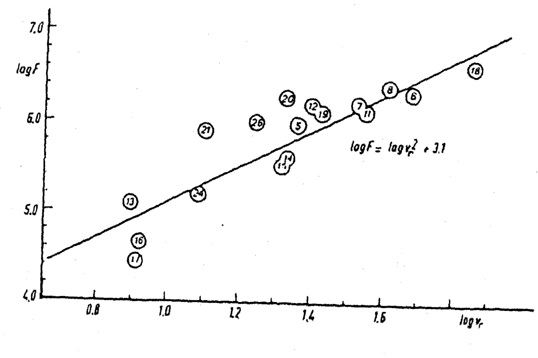

Figure 5

(Numbers refer to Table II)

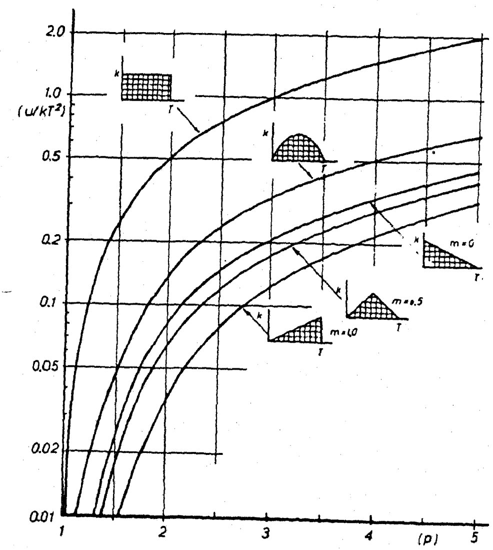

Figure 6.